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SOME IMPLICATIONS OF A DIRECT
MEASUREMENT OF THE HYDROGEN AND HELIUM ION
DISTRIBUTION IN THE UPPER ATMOSPHERE

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A recent daytime measurement of ion composition in the altitude region from 300 km to 940 km by Taylor, Brace, Brinton, and Smith (1963) allows some inferences about the physical processes governing the distribution of the light ionic constituents H^+ and He^+ . Although the absolute values for the ion concentrations derived from the current-measurements of the ion mass spectrometer may be subject to some errors, useful information can nevertheless be obtained from these data. The observed proton distribution has already been used by Hanson, Patterson, and Degaonkar, (1963) to deduce a value for the rate coefficient of the charge exchange reaction between hydrogen and oxygen.

By using the ratio of the two observed ionic constituents, $n(H^+)/n(He^+)$, it is possible to deduce additional information on the physical processes and structure parameters of the upper atmosphere.

In Fig. 1 the ratio $n(H^+)/n(He^+)$ based on the experimental data of Taylor et al (1963) is plotted as a function of geopotential altitude. It is quite obvious that the altitude distribution of this ratio can be divided into three distinct regions of behavior. These can be explained as the result of the predominance of either chemical processes (charge exchange, photoionization and ion-atom interchange or radiative recombination) or diffusion processes, or their combination as shown in the following:

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I. The lowermost portion of the distribution of $n(\text{H}^+)/n(\text{He}^+)$ in Fig. 1 can be explained by a chemical equilibrium distribution of the two ionic constituents.

It is now generally agreed that below 500 km the charge exchange reaction between hydrogen and oxygen governs the proton distribution (Hanson et al, 1963). Thus,

$$n(\text{H}^+) = \frac{9}{8} \frac{n(\text{H})}{n(\text{O})} n(\text{O}^+) \quad (1)$$

where $n(\text{X})$ is the number density of the constituent X, which is distributed exponentially with geopotential altitude according to its scale height, $H(\text{X})$.

The photochemical equilibrium distribution of He^+ is governed by

$$n(\text{He}^+) = \frac{I(\text{He})n(\text{He})}{K n(\text{O}_2)} \quad (2)$$

where $I(\text{He})$ is the photoionization rate coefficient for helium (Hinteregger and Watanabe, 1962) and K is the rate coefficient for the ion-atom interchange reaction with molecular oxygen, now considered to be responsible for the loss of helium ions (Bates and Patterson, 1962; Hanson, 1963).

It is easily shown from (1) and (2), that for the case of chemical and thermal equilibrium the ratio of the two ionic constituents follows the altitude dependence:

$$\frac{n(\text{H}^+)}{n(\text{He}^+)} = A \exp \left[-\frac{z}{H(2I)} \right] \quad (3)$$

where A is a constant, z is the geopotential altitude above the chosen reference level and $H(21)$ is the scale height corresponding to an effective mass of 21 A.M.U. (It is interesting to note that the altitude dependence given in (3) also holds even if the He^+ and H^+ distributions were both controlled by photochemical equilibrium, i.e., H^+ being produced by photoionization and lost by radiative recombination, assuming O^+ to be the predominant ion). Making use of (3) for the lowermost portion of the curve in Fig. 1 leads to an upper limit for the neutral gas temperature of 1200^{+60}_{-40}K which is in good agreement with the effective charged particle temperature $(T_e + T_i)/2 = 1235^{+40}_{-40}\text{K}$ inferred by Taylor et al, (1963) from the slope of their total ion density distribution. (The estimated neutral gas temperature for the time of this measurement, based on satellite drag observations (Jacchia and Slowey, 1963), is 1050^{+60}_{-40}K).

II. The uppermost portion of the $n(\text{H}^+)/n(\text{He}^+)$ distribution corresponds to a diffusive equilibrium distribution.

It is now generally known (Mange, 1960) that (unlike the behavior of neutral constituents) the diffusive equilibrium distribution of minor ions is not independent of other ionic constituents because of the presence of an electrostatic field (proportional to the mean ionic mass and the electron and ion temperatures) which is the same for all ions in the mixture and counteracts the gravitational force.

The diffusive equilibrium distribution of an ion (X_i^+) is given by:

$$n(X_i^+) = n_o(X_i^+) \exp \left[- \int_{z_0}^z \left(\frac{1}{H(X_i^+)} - \frac{\tau}{1+\tau} \frac{m_i g_o}{kT_i} \right) dz \right] \quad (4)$$

where $\underline{H}(X_i^+) = \frac{kT_i}{m_i g_0}$, with k Boltzmann's constant, T_i the ion temperature, m_i the ion mass and g_0 the acceleration of gravity at the earth's surface and where $\tau = T_e/T_i$ is the ratio of electron to ion temperature, z is the geopotential altitude and $m_+ = \Sigma n(X_i^+) m_i / \Sigma n(X_i^+)$ is the mean ionic mass.

It is easily seen from (4) that the ratio of two ionic constituents in diffusive equilibrium depends on the ion temperature only, even if there is absence of thermal equilibrium. For the two ionic constituents H^+ and He^+ this ratio can be expressed by

$$\frac{n(H^+)}{n(He^+)} = B \exp \left[\frac{z}{\underline{H}(3)} \right] \quad (5)$$

where B is a constant and $\underline{H}(3)$ corresponds to the scale height for an effective mass of 3 A.M.U. Using (5), an ion temperature $T_i = 1300^\circ \pm 60^\circ K$ is inferred from the top portion of the curve in Fig. 1, in good agreement with the temperature calculated for the chemical equilibrium region and with the effective charged particle temperature inferred by Taylor et al (1963).

III. The intermediate region in Fig. 1 corresponds to a variable scale height which is a weighted mean of the scale heights for the diffusive and chemical equilibrium regions.

As the result of the foregoing discussion the ion composition data of Taylor et al, 1963, can be interpreted in the following fashion:

Helium and hydrogen ions are in chemical equilibrium up to a geometric altitude of about 500 km. From this altitude to about 750 km, both chemical and diffusion processes seem to be influencing the distribution, while above 750 km the light ionic constituents appear to follow essentially a diffusive

equilibrium distribution.) It should be understood that departures from a true diffusive equilibrium distribution evidenced in the intermediate region may also arise from the effect of an upward or downward flux of ions as shown by Hanson (1963). The conclusion of diffusive equilibrium for the light ions above 750 km is difficult to reconcile with the observed decrease of $n(\text{He}^+)$ at these altitudes, unless one is willing to accept that the absolute values of the light ion concentrations reported by Taylor et al, (1963), are too low, or their values of total ion density (and $n(\text{O}^+)$ which is inferred therefrom) are too high, or that a combination of these factors is present. If the light ions were of comparable order of magnitude with $n(\text{O}^+)$ at altitudes above 750 km, $n(\text{He}^+)$ would indeed decrease with altitude as shown by Taylor et al, (1963), even though it is in diffusive equilibrium as indicated above, without reaching a second peak in its distribution. This fact can easily be verified from equation (4) by using a mean ionic mass appropriate to this situation. However, the magnitude of the adjustment of the experimental data of Taylor et al, (1963), required to produce the above effect does not appear to be justifiable on the basis of the internal consistency of their data (Taylor and Brace, private communication.)

The above conclusions, if supported by future measurements, may seriously impede efforts of explaining charged particle profiles in terms of a ternary ion-mixture model in diffusive equilibrium.

It is also quite evident that even measurements of relative ion abundances can provide useful information on the physical processes in the upper atmosphere. In particular, they can be used to determine ion temperature, up to now a parameter most elusive to direct measurements.

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Caption for Figure 1

Altitude distribution of ion concentration ratio $n(\text{H}^+)/n(\text{He}^+)$ derived from the ion mass-spectrometer measurements of Taylor et. al., (1963). The three regions indicated in the figure are interpreted in the text as corresponding to chemical equilibrium (I), diffusive equilibrium (II), and a transition from chemical to diffusive control of the distribution of H^+ and He^+ (III).

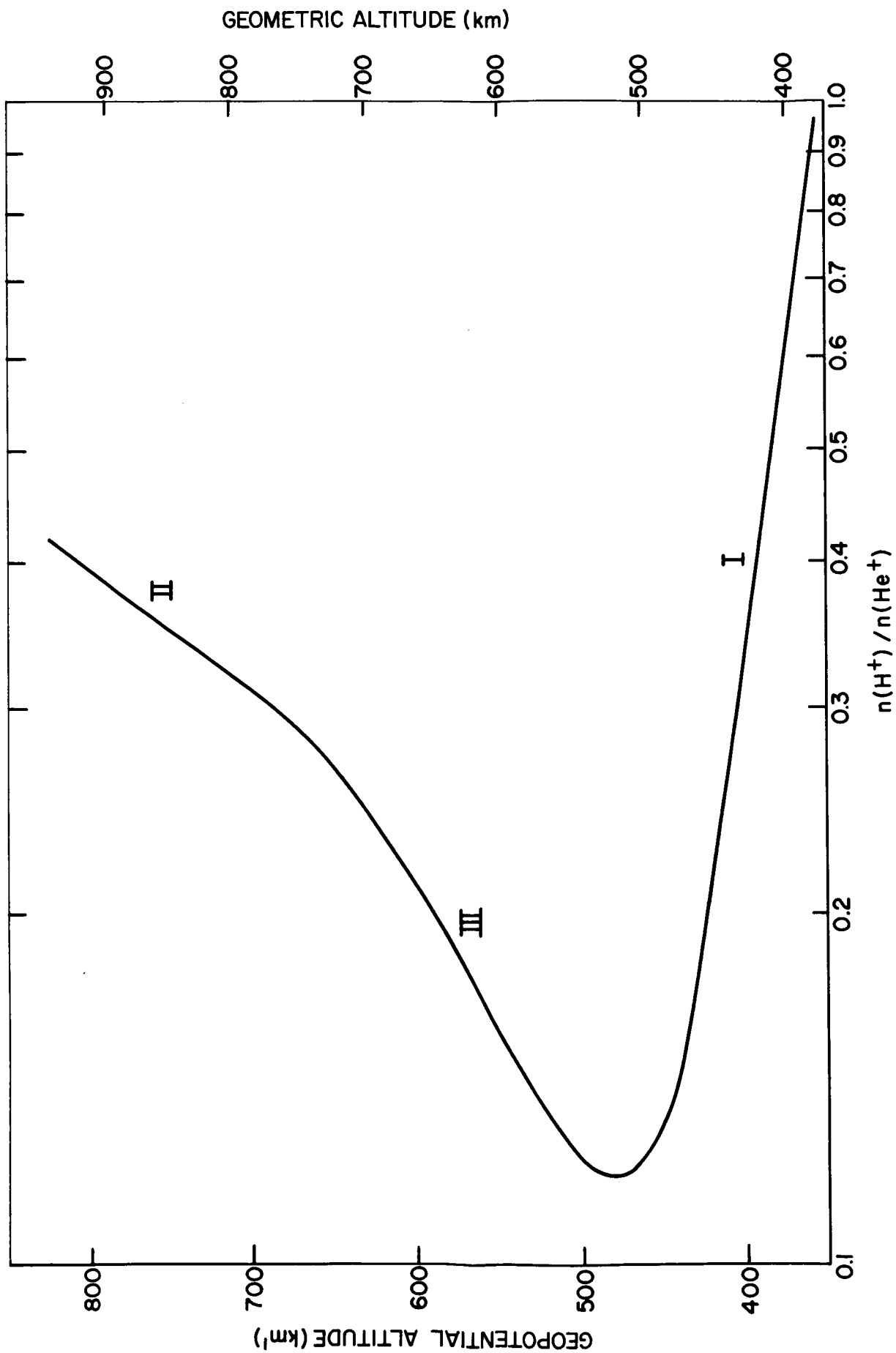


Figure 1